High-resolution morpho-tectonic profiling across an orogen

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6. Synthesis
"Orogenic stages and driving mechanisms"

Evolution of an orogenic convergent system can be described by a model of crustal accretion associated with plate subduction (Willette and Brandon, 2002). Crustal deformation is driven by a mass flux of sediments from the subducting plate into the orogen. During the convergent orogeny, a positive accretionary flux leads to surface uplift and an increase in erosional flux. The amounts of eroded material across the orogen and deposited in the adjacent basins help to derive the mechanical model of the subduction-collision tectonics of an Alpine-type compressional orogen (Beaumont et al., 1996). The eroded material can be controlled by parameters such as tectonics, surface processes, climate, topographic relief, etc. (Willett et al., 2006; Reiners and Brandon, 2006). Detrital thermochronology is a powerful tool to quantify erosion during the orogenic evolution (Rahl et al., 2007).

The process of orogeny in the SE Carpathians is described by a progressive accretionary wedge associated with inherent plate subduction asymmetry, which drives the shortening between the Tisza-Dacia block and the continental passive margin of the Moesian platform, from late Early Cretaceous to Present (details in Chapter 5 and summarized in Fig. 6.1). The material accreted from the westward advancing Moesian platform consists mainly of sediments, which were progressively eroded from the rising topography of the SE Carpathians and subsequently deposited in front of the advancing thrusting to the east. Orogenic stages in the SE Carpathians can be approximated based on the amounts of exhumation/erosion, which have been achieved by means of geomorphology and luminescence dating (Chapters 2 and 4) and low-T detrital thermochronology integrated with tectonics (Chapter 5), respectively. Because these methods address to different exhumation ranges, a problem occurred in integration of the thermochronological results generally sensitive to more than 1-2 km of exhumation (Ehlers and Farley, 2003) with the geomorphological markers (Burbank et al., 1996; Burbank, 2002) which are sensitive in the range of tens-hundreds of meters. In the absence of cosmogenic nuclides determination (no exposed surfaces), the link between the two methodologies was build assuming a linear interpolation.

Three stages have been distinguished: (1) subduction stage related to progressive growth of a thin-skinned thrust wedge, partly below sea-level, from Hauterivian? to Earliest Chattian (136-28 Ma), resulting in emplacement of the internal nappes; (2) subduction to collision stage with continuous growth of the accretionary wedge during the Earliest Chattian-Late Sarmatian (28-11 Ma), along with the emplacement of the external nappes and continental collision during the Late Sarmatian and (3) the Late Sarmatian-Holocene post-collision stage is characterized by high-angle reverse faults that affected the lower basement of the external nappes (11-0 Ma; Figs. 6.2 and 6.3).

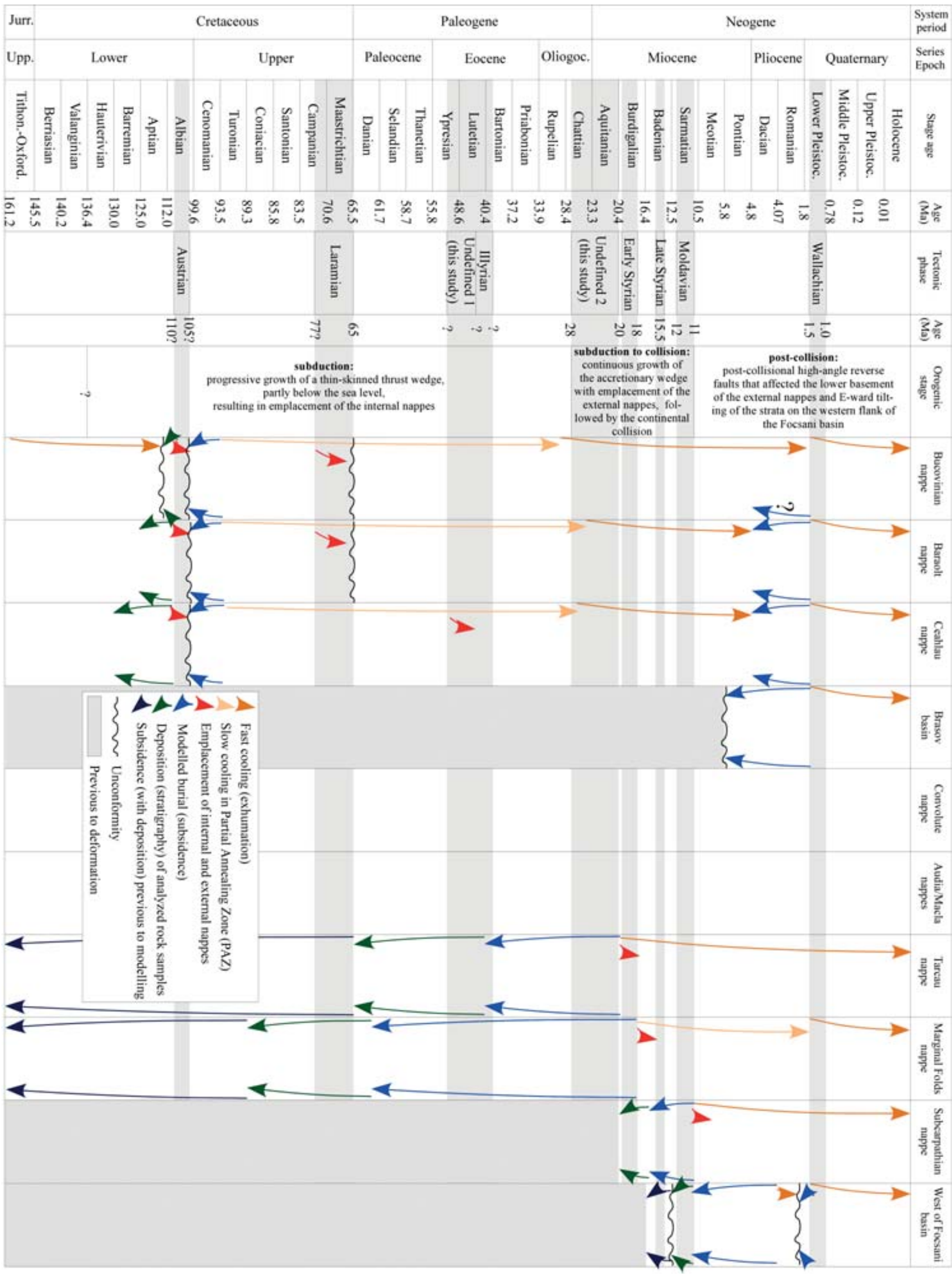
6.1. Hauterivian?-Earliest Chattian subduction stage (136-28 Ma)

A simple model for the thin-skinned thrust wedge is the critical taper (moving bulldozer; Dahlen, 1990), which assumes that subduction takes place against a rigid back-stop. If the convergence continues, the material accumulates and piles up internally until the surface slope reaches a certain angle, when the wedge slides along a décollment plane and begin accreting material at the toe. As the deformation takes place progressively as foreland-vergent thrusting, it leads to topographic growth.

For the SE Carpathians, the Upper Jurassic-Lower Cretaceous distal oceanic sediments and slope turbidites have been scraped off in an accretionary wedge during the late Early Cretaceous, which was the first contractional event recorded in the Ceahlău-Severin oceanic domain. This is usually interpreted as the onset of subduction in this domain (Săndulescu, 1988), which was probably due to deep mantle processes such as slab detachment (Wortel and Spakman, 2000), slab pull (Martin et al., 2006) or delamination (Sacks and Secor, 1990). It resulted in stacking of the internal nappes, with peak deformation in intra-Albian (Austrian phase; Fig. 6.1). Subsequently, the internal nappes underwent burial (~3.2±0.5 km) during the Late Albian-Cenomanian, which continued also during the entire Paleogene, attaining a maximum of 4.2±0.7 km in the east (Figs. 6.2a and 6.3a). The internal nappes were affected by two shortening events with rather limited exhumation, during the Late Campanian-Maastrichtian and Eocene, respectively (Fig. 6.1). While the first phase is known in the local literature ("Laramian phase", Săndulescu, 1988), the second one was not defined so far. However, the latter phase was assumed to exist (Schmid et al., 2008; see also Chapter 5) based on coeval deformation observed in the Balkans (final continental collision with the Moesian platform), which peaked during the Middle Eocene (Illyrian phase; Ivanov, 1988; Bergerat and Pironkov, 1994; Fig. 6.1) and exhumation in the South Carpathians during the Paleogene (Fügenschuh and Schmid, 2005).

The central and eastern internal nappes experienced exhumation of 1.4±1.7 km, and therefore, this part of the orogen was already under erosional conditions during the Late Cretaceous-Paleogene shortening. However, the nappes remained at few kilometres deep until the Earliest Chattian. This suggests that the exhumation was much smaller than the subsidence to cause subaerial exposure of the nappes. However, more to east in the foreland basin, the subsidence was faster than exhumation (Fig. 6.2a), the nappes remaining under the sea level as demonstrated for instance by the sedimentological character of the Upper Campanian-Maastrictian post-kinematic cover (Melinte and Jipa, 2005). Erosion affecting the internal part of the thin-skinned wedge provided sedimentary influx for the adjacent basins (Fig. 6.3a), subsequently, the nappes being covered by an unconformable Paleogene turbiditic sequence of variable thickness and lithology (Fig. 6.1). The high subsidence experienced by the internal and external nappes during the Late Albian-Earliest Chattian might suggest that the subduction in the SE Carpathians was dominated by a slab-driven mechanism (Vrancea slab), such as the slab-pull and/or roll-back of the Carpathians embayment (Royden, 1993). Such a deep-seated mechanism would prevent "normal" (or better said simple) orogenic exhumation during this stage.

Figure 6.1



Stratigraphic correlation and tectonic evolution scheme of the SE Carpathians from Lower Cretaceous to Quaternary. Tectonic phases are: Austrian (intra-Albian), Laramian (Late Campanian-Maastrichtian), Undefined 1 (Ypresian-Lutetian) equivalent? to Illyrian (Middle Eocene in Balkans), Undefined 2 (Chattian-Aquitanian), Early Styrian (intra-Burdigalian), Late Styrian (intra-Badenian), Moldavian (Late Sarmatian) and Wallachian (?late Early Pleistocene). Tectonic phases are mainly based on Săndulescu (1988), Illyrian phase is from Ivanov (1988) and Bergerat and Pironkov (1994) and Undefined 1 and 2 are based on the results from this study.

6.2. Earliest Chattian-Late Sarmatian subduction to collision stage (28-11 Ma)

Prior/or close to the Late Oligocene (Earliest Chattian), the internal nappes were sutured in a single block, herein defined as the internal indentor, which later acted as a rigid back-stop against the westward subduction of the distal parts of the Moesian continental passive margin (Fig. 6.3b). The renewed convergence between the internal intendor and the Moesian platform triggered by the clockwise rotation of the Tisza-Dacia block (Füegenschuh and Schmid, 2005) has caused exhumation in the internal nappes and progressive deformation in the external nappes, the eastern part of the thin-skinned wedge. The exhumation recorded in the internal intendor attained a maximum of 2.2±0.6 km, slowly decreasing eastward (Fig. 6.2b).

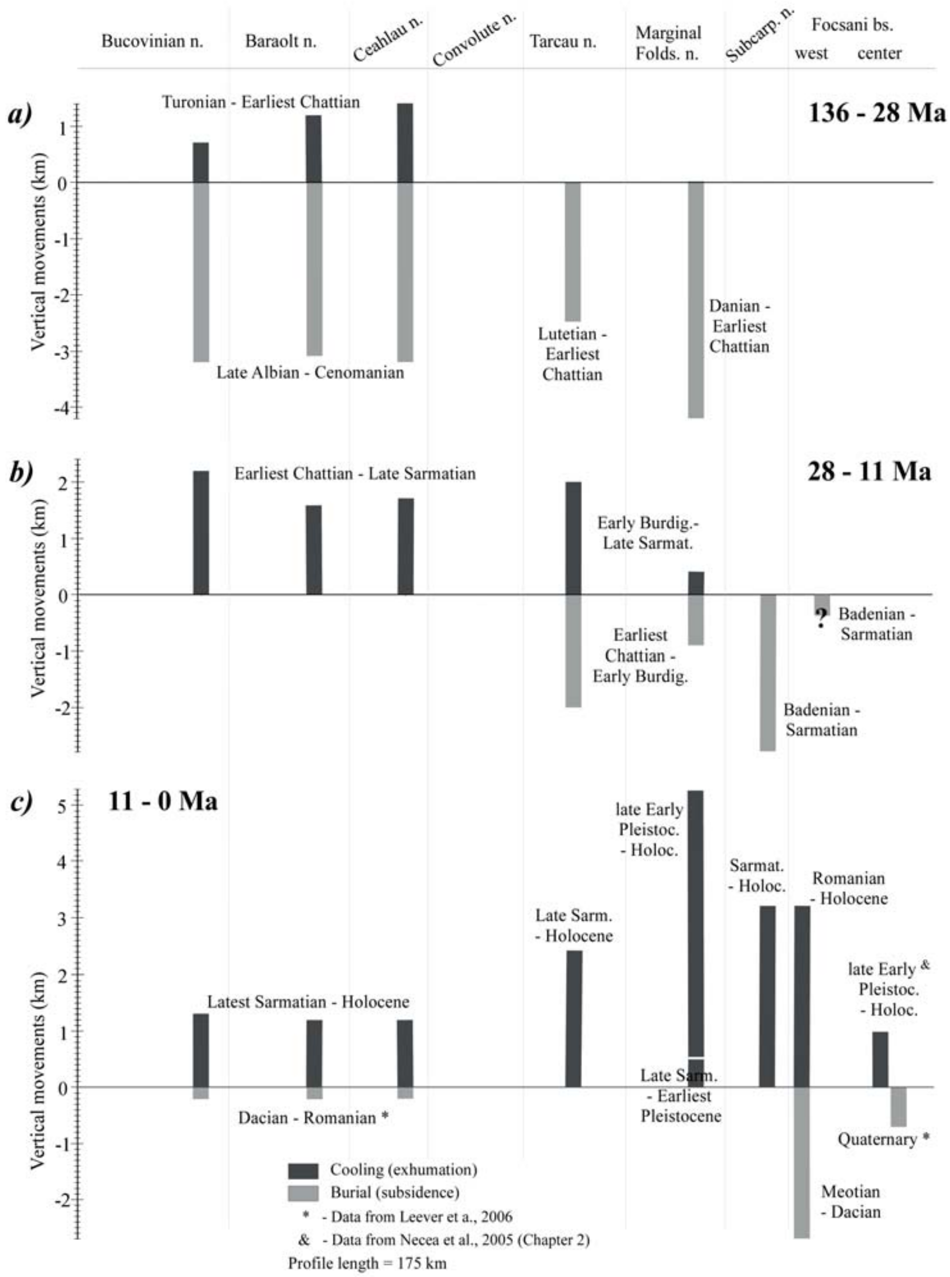
In the east, shortening/thrusting took place during the entire Miocene, resulting in the gradual emplacement of the external nappes (Early and Late Styrian phases), with the final docking against the Moesian platform during the Sarmatian (Moldavian phase; Fig. 6.1). The gradual nappe emplacement towards the foreland basin resulted in flexural subsidence, the depocentre being gradually shifted to the east (Fig. 6.3b). The exhumation of the western external nappe is ~2.0±0.3 km, value which decreased gradually eastward towards the Carpathian foreland (Fig. 6.2b). This gradual exhumation of the thin-skinned thrust wedge provided source material for the Carpathian foreland and the eastern part of the Transylvanian basin (Fig. 6.3b). The subsidence recorded during the Earliest Chattian-Late Sarmatian reached a maximum of 2.8±0.4 km in the most external nappe (Fig. 6.2b). The Middle Miocene (Badenian)-Sarmatian subsidence that affected the central and northern parts of the foreland basin, outside of the study area, reached a higher value of ~4-5 km for non-decompacted sediments (Tărăpoancă et al., 2003).

An overall amount of 130-180 km of Miocene shortening was recorded by the East Carpathian thin-skinned units during the complex mechanism of rotation, translation and docking of the Tisza-Dacia internal units against the European/Moesian stable foreland (Ustaszewski et al., 2008). The coeval exhumation recorded in the order of less than ~3 km is rather low when compared with other orogens such as the Alps, where collisional exhumation is in the order of tens of kilometres (Schmid et al., 1996). Moreover, this exhumation amount is by far much less than that one predicted by modelling the exhumation of simple orogenic wedges (Beaumont et al., 1991; Hoth et al., 2007). Therefore, these data are in agreement with the earlier inferences on subduction mechanics, which suggest that the subduction and collision in the Carpathians was still dominated by the earlier slab-driven mechanism. The Earliest Chatthian to Late Sarmatian period corresponds to transition from subduction to collision. During the Late Sarmatian (~11 Ma), continental collision took place between the most external nappe and the Moesian platform.

6.3. Late Sarmatian-Holocene post-collision stage (11-0 Ma)

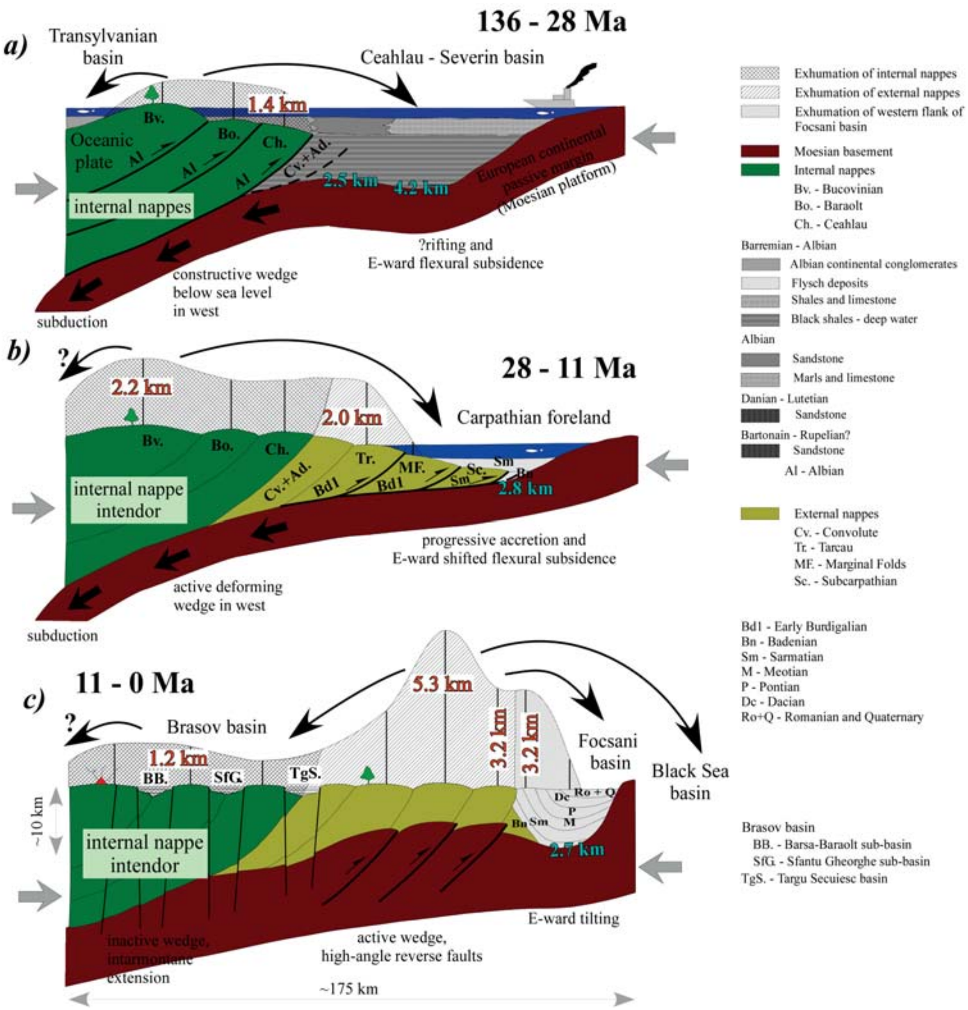
At the local SE Carpathians scale, the contraction recorded during this stage is due to the post-collisional renewed convergence between the SE Carpathian orogen and the Moesian platform, resulting in development of high-angle reverse faults located in the lower basement of the external nappes (Fig. 6.3c; Bocin et al., 2005; Leever et al., 2006). Exhumation affecting the external nappes during this stage has an asymmetric pattern,

Figure 6.2



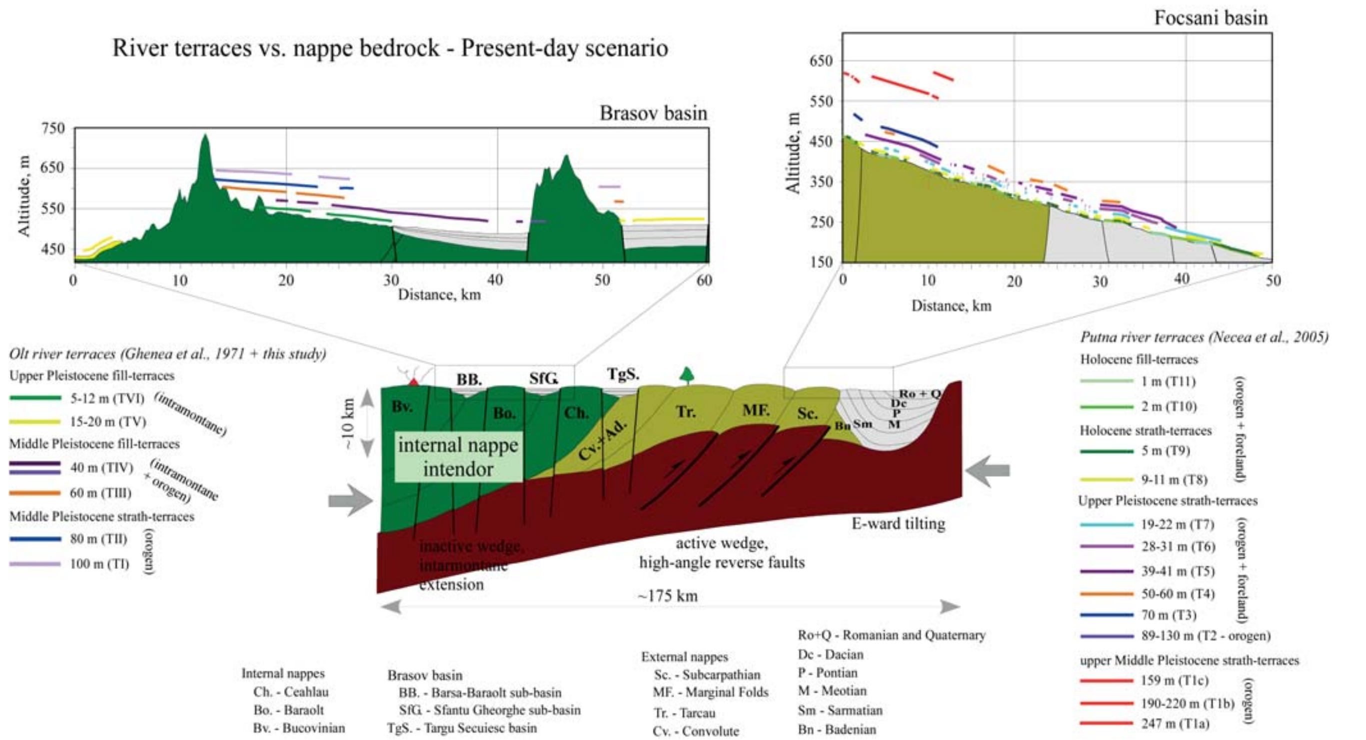
Cumulative burial/exhumation (erosion) amounts quantified across the SE Carpathians and adjacent Focşani basin, based on which have been separated three orogenic stages: (a) subduction from Hauterivian? to Earliest Chattian (130-28 Ma), subduction to collision from Earliest Chattian to Late Sarmatian (28-11 Ma) and (c) post-collision from Late Sarmatian to Holocene (11-0 Ma; explanation in text). Dark-grey colour stands for cooling (exhumation) and light-grey for burial (subsidence).

Figure 6.3



Three-stage orogenic evolution of the SE Carpathians. (a) Hauterivian?-Earliest Chattian subduction stage: progressive growth of an accretionary wedge (dark- and light-green colours) below the sea level with emplacement of the internal nappes and ?rifting and E-ward flexural subsidence (130-28 Ma); (b) Earliest Chattian-Late Sarmatian subduction to collision: active deforming wedge in the west, continuous growth of the accretionary wedge in the east with emplacement of the external nappes (light-green colours) and E-ward shifted flexural subsidence, followed by the continental collision during the Late Sarmatian (28-11 Ma) and (c) Late Sarmatian-Holocene post-collision stage: inactive wedge in the internal nappe intendor, post-collisional high-angle reverse faults that affected the lower Moesian basement (brown colour) of the external nappes and E-ward tilting of the Upper Sarmatian-Lowermost Pleistocene strata of the Focşani basin (11-0 Ma). Exhumation is represented by different patterns in the internal and external nappes and on the western flank of the Focşani basin. Red values are examples of exhumation amounts, while light-blue values are examples of subsidence amounts. The horizontal and vertical scales refer to the present-day scenario. Figure inspired after Sanders et al., 1999.

Figure 6.4



River terraces (from Chapter 4) plotted along the geological cross-section (from Chapter 5). Notice that the oldest river terraces, Middle Pleistocene in age, cover mainly the orogenic nappes (description in text).

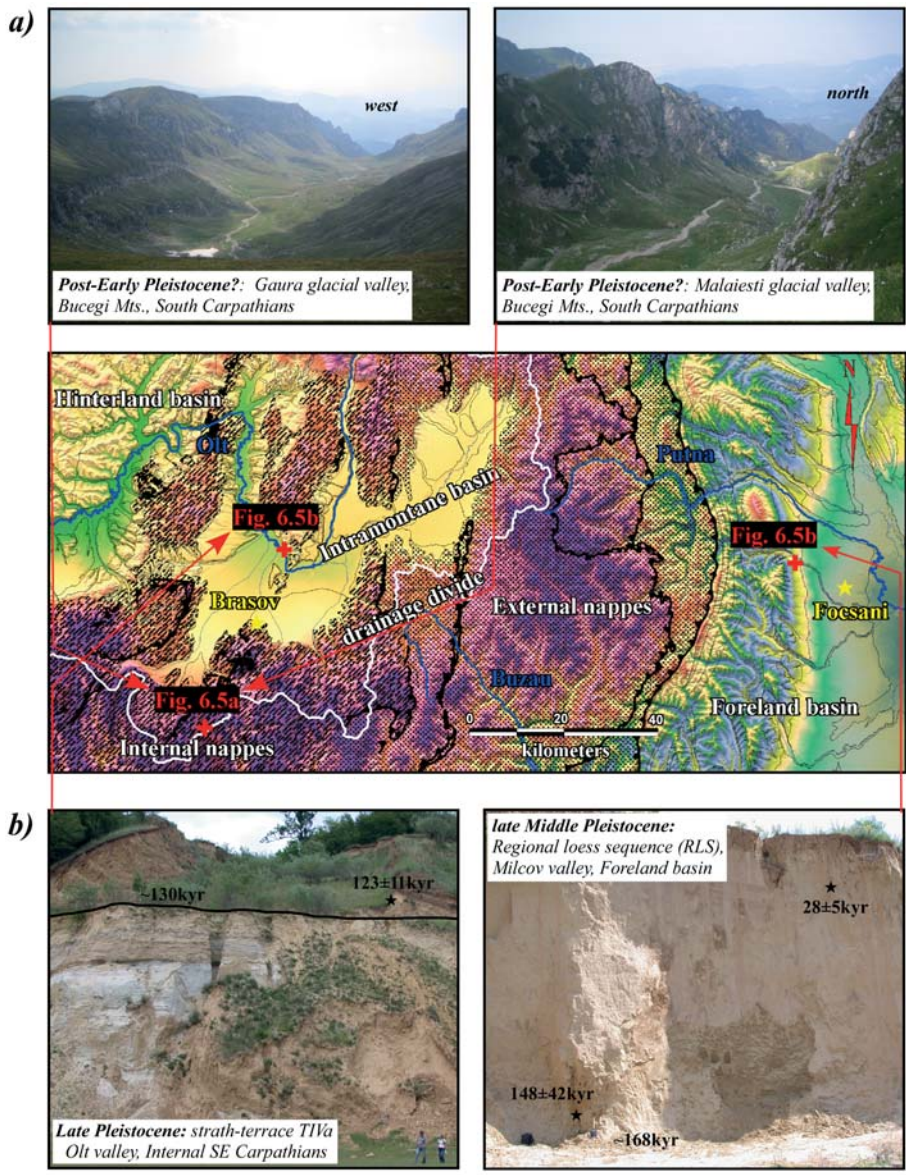
reaching its maximum during the late Early Pleistocene in the central external nappe (Wallachian phase; Fig. 6.1). The overall material removed from the nappes is ~1.2±0.5 km in the internal intendor, increasing up to 5.3±0.7 km in the external nappes and decreasing further eastward to 3.2±0.5 km on the western flank of the Focşani basin (Fig. 6.2c). The coeval subsidence was restricted eastward, attaining a value of 2.7±0.4 km (Fig. 6.2c). However, the maximum subsidence that took place during the Late Sarmatian-Quaternary was recorded further to the SE in the basin centre, outside of the study area (~7-8 km; Tărăpoancă et al., 2003).

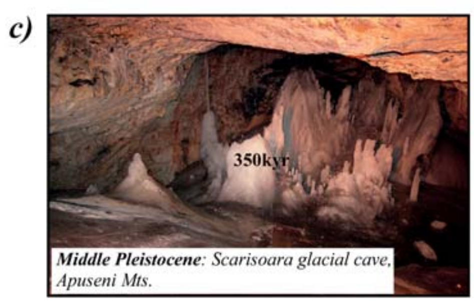
The large sedimentary influx provided by the exhuming nappes was deposited locally in the Focşani and Braşov basins, and probably larger amounts being transported and deposited by the Danube river in the Black Sea basin (Fig. 6.3c). In this basin, a striking increase of sedimentary influx during the Latest Miocene-Quaternary is recorded in the Danube discharge area, coeval with inferred moments of complete basin fill (or complete basin drain) in the Focşani area (Dinu et al., 2005; Gillet et al., 2007). During this stage, the SE Carpathians underwent post-collisional continuous exhumation from Late Sarmatian to Present (Figs. 6.2c and 6.3c), being more accelerated since the late Early Pleistocene (4.8±0.4 km; Fig. 6.2c).

Whether or not this is the effect of a Pliocene-Quaternary inversion recorded at the larger Pannonian-Carpathians scale (Matenco et al., 2007; Heidbach et al., 2007), it is clear that deformation in the orogen has radically changed its mechanics. The 5 km Pliocene-Quaternary shortening (Leever et al., 2006) accommodated by ~5 km exhumation is in striking contrast with the 130-180 km of shortening accommodated by ~3 km exhumation. At the crustal level, the amount of exhumation equals the amount of crustal thinning observed below the SE Carpathians orogen (Hauser et al., 2007), which is rather compatible with the large scale folding speculated (Leever et al., 2006; Matenco et al., 2007) and confirmed by the observed crustal geometries (Schmid et al., 2008, in plate 3 section East Carpathians). The exhumation of the SE Carpathians is mainly attributed to the high-angle reverse faults that affected the basement underlying the external nappes. The contraction during this stage might be related to the Vrancea slab dynamics that operated in the aftermath of the Late Miocene continental collision (Matenco et al., 2007).

The late Early Pleistocene exhumation of the central external nappe was responsible for the ENE-ward tilting of the Upper Sarmatian-Lowermost Pleistocene strata from the Focşani basin (85° to 9°). The response of the Carpathian river network to this deformation was recorded as a stair of river terraces, from Middle Pleistocene to Holocene, which were progressively uplifted and locally tilted to the east (see Chapters 2 and 4). Figure 6.4 shows the relationship between the river terraces and the underlying basement represented by the orogenic nappes. In the west, the older Middle Pleistocene river terraces (levels TI-TIII; 100 to 60 m) indicate maximum uplift in the internal nappes, while the late Middle-Late Pleistocene levels, e.g. TIV-TVI (40 to 5 m) equally covering the internal nappes and the sedimentary filling of the intramontane basin point to a more generalized deformation. To the east, the late Middle Pleistocene river terraces, levels T1-T3 (250 to 70 m) overlap the central external nappes, an area which corresponds to the late Early Pleistocene exhumation shown by the thermochronological data. The younger terrace levels, Upper Pleistocene-Holocene ages (T4 to T11; 60 to 5 m) extend eastward, progressively covering the orogen-foreland transition zone and the foreland itself, which indicate an eastward migration of deformation from Middle Pleistocene to Holocene.

Figure 6.5





Evidences for climatic conditions recorded in the internal and external Carpathian nappes. (a) Post-Early Pleistocene? glacial valleys located in the internal zone of the South Carpathians, i.e. Gaura and Mălăieşti U-shaped valleys in the Bucegi Mountains; (b) late Middle Pleistocene and Late Pleistocene regional loess sequences dated by IRSL that cover large areas in the foreland basin and the internal orogenic units; (c) Middle Pleistocene Scărişoara glacial cave located in the Apuseni Mountains (http://www.flickr.com/photos/tsk/2598037525/in/photostream) was dated through U-series by Onac in 2001.

The abrupt change in the erosional pattern recorded across the orogen during the late Early Pleistocene-Holocene period lead, among others, to deposition of the conglomerates/gravels with an unprecedented coarse character for the entire Miocene collisional evolution. This character is justified by the radical change in the exhumation velocities, but a coeval significant change in the climate cannot be excluded. The climatic conditions are sustained by the glacier traces found as U-shaped valleys with basal moraines in the alpine zone of the South Carpathians and by the loess deposits largely preserved in the hinterland and foreland basins (see Chapter 4). The glacial valleys identified in the internal part of the Carpathian curvature are N-S to W-E-orientated, suspended at altitudes higher than 1700 m (Fig. 6.5a; Mălăieşti and Gaura valleys, Bucegi Mts., South Carpathians). This denotes that the mountains acted as a barrier against the advancing glaciers, which developed at altitudes of 2500-1700 m only in the internal part of the mountain chain. The previous geomorphological studies indicate that the Lowermost Pleistocene deposits are climatic-related in the eastern part of the South Carpathians (Bucegi Mountains; Ghenea, 1970; Săndulescu, 1984), while other sources point to a lacustrine environment (Jipa, 1997). The latter hypothesis is more plausible because it can be sustained by the regional palinspastic reconstructions, which indicate an Earliest Pleistocene water connection between the Braşov intramontane and Focşani foreland basins (Leever et al., 2006). Further to the west, the moraines dated by ^{10}Be exposure and pedological investigations indicate an extensive Late Pleistocene glaciation (Reuther et al., 2007). The loess sequences identified on both sides of the orogen (Fig. 6.5b) and dated by luminescence (Chapter 4) indicate that the loess accumulated during several cold stages. The older one corresponds to the Riss glacial (late Middle Pleistocene) and is located eastward in the foreland basin (Fig. 6.5b). The glacial remnant preserved in the Scărişoara cave (Fig. 6.5c; Apuseni Mountains) has an age of 350 kyr based on the U-series analysis performed on stalagmites and flowstone fragments, which indicates that the glacier formed during the Middle Pleistocene (Onac, 2001). From the above data, the onset of glaciation in the Romanian Carpathians can be roughly placed during the Middle Pleistocene, mainly overlapping the internal nappes.

Formation of terrace levels TI to TVI (internal nappes) and T1 to T11 (external nappes) during either full climate (i.e. warm or cold) or climatic transitions indicate that the river incision did not fit preferentially to warm, cold or climatic transition. This suggests that both tectonic and climatic processes could have controlled the river incision and terrace formation in the SE Carpathians. In the external nappes, the river terraces are directly controlled by the high-angle reverse faults that affected the lower basement of the nappes. During the Middle Pleistocene-Holocene, the Black Sea level dropped twice with 100 and 60 to 70 m, respectively, which might have played only a minor role on river incision and terrace formation.